DYNAMICAL CLIMATOLOGY TECHNICAL NOTE NO. 85

SIMULATION OF SEA-ICE IN A COUPLED OCEAN-ATMOSPHERE MODEL

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December 1989

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1. INTRODUCTION

Sea-ice plays a unique and important role in the global climate system. It not only interacts with both the ocean and the atmosphere, but interferes with direct communication between them. In attempting to simulate climate by developing global coupled atmosphere-ocean models, we are thus obliged to devote considerable attention to modelling sea-ice.

This paper describes results from a run of the UK Meteorological Office coupled atmosphere/ocean/sea-ice model. Although the total length of the experiment was almost four years of simulation time, several improvements were made to the sea-ice model as the run progressed. The results to be shown here are thus taken from the second half of the experiment, after the formulation of the model had crystallised. Of course, a far longer integration would be needed for the model to reach equilibrium, but much can be learnt about the processes in the model without knowledge of its equilibrium state (or states). Note that the scope of the discussion will not be limited to ice. This is because the highly interactive nature of the climate system makes it very difficult to understand the behaviour of one component, such as sea-ice, in isolation from the others. Furthermore, from the viewpoint of climate modelling it is actually the realism with which the entire system is simulated that is the paramount consideration. Thus the approach taken here will be to consider sea-ice in the context of the coupled model as a whole. However, before the results are described we consider some relevant features of the coupled model system, in particular the sea-ice model itself.

2. THE COUPLED MODEL FORMULATION

The component models employ a 2.5° x 3.75° latitude-longitude grid, with 11 layers in the atmosphere and up to 17 in the ocean. Instead of exchanging information continually, the ocean and atmosphere models run separately, and update each other's surface boundary conditions at 5 day intervals. However, the sea-ice model is split between the ocean and atmosphere parts of the system, in such a way that ice surface temperatures change during the atmospheric segment, and ice depths and extents are updated during the oceanic segment. As explained by Gordon & Bottomley (1984), this division is appropriate because ice formation must occur in the ocean model, while the diurnal cycle of ice surface temperature can only be properly represented in the atmosphere model. A disadvantage of the scheme is that the sea-ice depths and extents seen

by the atmosphere only change every 5 days. It is therefore likely that future experiments will use a shorter coupling interval, such as 1 day.

The atmosphere model is the version of the Met O 20 climate model used in the so-called '4th Annual-Cycle' (4AC) benchmark experiment, described by Slingo *et al.* (1989). Unlike earlier versions, this one calculates the radiative fluxes using the simulated cloud distribution rather than prespecified climatological cloud amounts. In principle the removal of this constraint constitutes an important step forward in the development of the model, but it also means that there is more scope for the radiative fluxes to deviate from realistic values. It should be noted too that the radiative properties of the modelled clouds are assigned constant values intended to be representative of the whole globe. A modelled cloud in polar regions is therefore optically thicker than a cloud of similar depth would be in reality. (See, for example, the article on Arctic stratus clouds by Herman (1986).)

The ocean model is a derivative of the GFDL model described by Cox (1984), incorporating the isopycnal diffusion scheme of Redi (1982), and the stability-dependent vertical diffusion scheme of Pacanowski & Philander (1981). Because of the importance for a coupled model of achieving a good simulation of SST, the vertical resolution is greatest near the surface: the top 4 layers are each 10m thick. The velocity, temperature and salinity fields are Fourier filtered polewards of 75°N, to prevent instabilities due to the small zonal grid spacing there. As it was anticipated that filtering the temperature field could artificially initiate ice formation, this is prevented by invoking the filter later in the timestep than the test for freezing.

The sea-ice model is based on the 'zero-layer' thermodynamic model of Semtner (1976), in which the prognostic variables are surface temperature, ice depth, and snow depth. Temperatures within the ice and snow are not computed; the internal temperature field is assumed to be piecewise linear, with a change of gradient at the ice/snow interface because of the lower thermal conductivity of snow. In order to simulate the diurnal cycle of surface temperature as realistically as possible within the framework of the zero-layer model, a surface heat capacity term is introduced, as described by Gordon & Bottomley (1984). Although there is no representation of ice dynamics, Semtner's model has been modified to allow the presence of leads, by introducing ice concentration, A, as another prognostic variable. Changes in A due to thermodynamic processes are modelled using a formulation due to Hibler (1979), which assumes that the thicknesses of ice floes in a grid box are uniformly distributed over the interval (0, 2h/A), where h is the mean ice depth for the entire box. Since completely unbroken ice cover is very rarely encountered in reality, because of small-scale deformations of the ice-pack, A is prevented from exceeding 0.995 in the Arctic and 0.98 in the Antarctic. This practice was adopted by Parkinson & Washington (1979); the limits are somewhat arbitrary but reflect observational evidence that Antarctic ice tends to be less compact.

The presence of ice adds a number of extra processes to the system, as well as modifying existing ones. The representation of these will now be described, beginning with the atmosphere/ice interactions.

The albedo of the snow/sea-ice surface, α, is specified as a function of surface temperature, T, by fixing $\alpha = 0.8$ at and below T = 268K, $\alpha(273) = 0.5$, and varying α linearly between these two points. This is a simple parametrization of the effect of melt-ponds and puddles. No account is taken of the ice thickness, the depth of snow cover, or the increase in (spectrally-averaged) albedo under overcast skies. (The importance of these and other effects is discussed in a paper by Shine & Henderson-Sellers (1985).) However the effect of multiple reflections between the snow/ice surface and the base of cloud layers is dealt with, by making an appropriate reduction in the surface albedo, as detailed by Slingo (1985). The albedo of leads is set to 0.06, the same value used for areas of ice-free ocean.

Because leads significantly enhance the turbulent heat fluxes in the atmospheric boundary layer, the standard (4AC) version of the boundary layer scheme is modified in the coupled model, so that the fluxes are calculated separately over the portions of a grid box covered by ice and by leads. The appropriate (bulk) Richardson number is used to characterise the stability of the surface layer in each case. Since the experimental investigation of boundary layer processes over sea-ice is still an area of active research, it is probable that the details of the current formulation could be improved upon. However it should still be an improvement on the position in the 4AC model, in which the effects of leads on the boundary layer are not represented at all.

There are 3 main ways in which sea-ice directly interacts with the ocean beneath it. One of these, momentum transfer, is not represented at all in the present model, because of the omission of ice dynamics; instead the computed atmospheric wind stress is passed directly to the ocean. The other two are heat and salt transfer: these are parametrized in simple ways which will now be outlined.

The heat flux from the ocean to the ice, F, is assumed to be related linearly to the temperature of the top layer of the ocean, T, and the temperature at the base of the ice, t, by $F = K(T - t)$. Here t is the freezing point of sea-water; its dependence on salinity is ignored and t set to -1.8°C everywhere. K is a tunable parameter, given the value 200 $Wm^{-2}K^{-1}$ in the model integration described below. This representation of F has been used before, e.g. by Pease (1975). It has been criticised on several grounds by Parkinson & Washington (1979), but is perhaps the best that can be done without implementing a proper boundary layer scheme.

In parametrizing the effect of sea-ice processes on the salinity of the ocean, the salinity of ice, s, is taken to be uniform. This is quite unrealistic, but justified in view of the complications involved in keeping track of the history of the ice. The value assigned to s is 0.6‰. When sea-water of salinity S freezes, salt is rejected in proportion to the excess (S - s), increasing the salinity of the surface layer of the ocean. When ice melts, the relatively fresh melt-water reduces the salinity of the surface layer.

3. RESULTS OF THE INTEGRATION

The models were initialised with (June) data from previous runs: the atmospheric data came from the 4AC integration, while the ocean data were produced by a two year run of the ocean model, starting from a state of rest with climatological temperature and salinity fields (Levitus, 1982), driven by surface fluxes from the 4AC integration with sea surface temperatures and surface salinities relaxed back towards their climatological values. This was intended to ensure that the initial state of the ocean would be both fairly realistic and compatible with the 4AC atmosphere. Similarly, sea-ice extents were taken from the climatology used in the 4AC run, with the thicknesses and concentrations initialised at 0.5m and 80% for seasonal ice, and 2.5m and 98% in permanently ice-covered areas.

After the run had finished it was discovered that, due to a programming error, the solar radiation absorbed by leads was effectively twice the correct amount. This error would tend to promote the melting of ice in summer, and should be taken into account when assessing the results.

The final year of the integration does not constitute an equilibrium cycle, as explained previously. However the model's behaviour is sufficiently consistent that the results to be shown are representative. Consider first Figures 1 and 2, which show the monthly average sea-ice thicknesses in the Arctic, during March and September of the final year of the experiment. In the central Arctic, the order of magnitude of the ice thicknesses is reasonable, although the greatest values occur near the Pole and east of Severnaya Zemlya. In reality, the thickest ice is found along the north coast of Greenland and near the Canadian Archipelago.

The overall extent of the ice cover may be compared with mean ice extents for 1973-76, shown in Figure 5. These were derived by Parkinson *et al.* (1987) from data obtained by the passive microwave radiometer on the Nimbus 5 satellite. One must remember, though, that Figures 1 and 2 show individual model years and that there have been significant changes in ice extents on both interannual and decadal timescales during this century, so that the comparison is not fully valid.

The amplitude and phase of the simulated seasonal variation in ice cover are reasonable, but it is clear that ice is too extensive throughout the year. For example, it persists during summer in Baffin Bay and the northern half of Hudson Bay, areas which should be ice-free by September. In winter, ice appears in the Gulf of Alaska and covers a wider area of the Greenland and Norwegian Seas than in reality, at least in recent history. Finally, the Barents Sea, which should not be fully ice-covered even in winter, remains icy all the year round (though with extensive leads in summer).

The situation is very different in the Antarctic, as can be seen from Figures 3 and 4, which show the monthly average ice thicknesses during February and October in the last year of the run. There is too little ice in October and too much in February, so that the model

completely fails to produce the large annual variation in ice extent that is found in reality. This is clear, for example, from the satellite data for 1973-76 given by Zwally *et al.* (1983), shown in Figure 6. (It should be noted that this data is known to be somewhat unrepresentative, because of the occurrence of the famous Weddell Sea polynya in the years 1974-76. This phenomenon did not recur during the period 1977-85, according to Washington & Parkinson (1986).) The total amount of ice also displays a gradually declining trend, at least over the second half of the integration. This is apparent in Figure 7, which shows the difference in the thickness of ice between the Februaries of year 3 and year 4. Although ice has grown in some areas near Antarctica, such as in the Ross Sea, it has been systematically eroded in regions adjacent to the open ocean.

The seasonal evolution of ice concentration is also of interest. In the central Arctic this appears to be fairly realistic: the ice cover remains almost complete until the end of May; leads appear during June and broaden in July; the ice concentration is a minimum in August at about 80% (see Figure 8); by the end of October the ice cover has been fully restored. In the Antarctic the picture is less clear because of the difficulties already mentioned.

As noted by Roberts (1988), the forcing of the salinity field through brine rejection during freezing or surface freshening during ice melt (section 2) has a marked modulating effect on the ocean-ice heat flux, F. In winter the rejection of salt destabilises the water column and promotes vertical mixing of heat, leading to large values of F in the seasonal ice zone (up to 100 W M^{-2}or more). During the melting season, the freshening of the surface layer inhibits vertical mixing and thus substantially reduces F. As implied, the seasonal differences are most pronounced in the seasonal ice zone on the periphery of the Arctic, because the rates of freezing and melting are largest there, and the ocean heat supply is replenished by advection. Values in the central Arctic are only of order a few watts per square metre, being slightly higher in winter than summer.

4. DISCUSSION

It has been shown that the simulation of sea-ice in the coupled model, though by no means completely unrealistic, has some significant deficiencies. We would naturally like to make improvements that would alleviate these or even eliminate them altogether. However, it is first necessary to decide which aspects of the model are most in need of improvement.

Consider the simulation in the Antarctic first. We believe that our failure to simulate the large annual variation in sea-ice extent is mainly due to the absence of ice dynamics from the model. The main evidence for this is a paper by Hibler (1984), in which he compared the performance of models of Antarctic sea-ice with and without dynamics, and found that the amplitude of the seasonal variation in ice area was larger and more realistic when dynamics was included. He also gave a convincing analysis of why this should occur, which may be summarised by

saying that the prevailing winds tend to induce a northward drift of the ice near the edge. This promotes the advance of the edge in winter, when the ice is transported into water that is cooling in any case, but accelerates decay in summer, since the ice is then transported into water that is tending to warm.

Another cause for concern is the tendency for the Antarctic ice cover to decline. This trend is reminiscent of problems encountered by previous investigators, e.g. Gates *et al.* (1985), who found that in the course of a 17-year experiment with their coupled model, the annual maximum in Antarctic sea-ice area diminished to about a quarter of its initial value. Similarly, Washington *et al.* (1980) reported that there was far too little Antarctic sea-ice in their coupled model. They suggested that this was mainly due to the excessively large horizontal diffusion of heat in the ocean, which in turn was a consequence of the large horizontal grid spacing (5° x 5°). This is also likely to have been a problem with the model of Gates *et al.* (1985), which used a 4° x 5° grid. However Gates *et al.* (1985) were more inclined to favour an explanation based on deficiencies of their atmospheric model, such as excessive solar radiation at the surface, caused by insufficient cloudiness. The present model has finer resolution (2.5° x 3.75°), but may suffer from high diffusion to some degree; the problem deserves further investigation.

A point worth considering is that the feeble seasonal cycle of ice extent mitigates this problem to some extent, as it must tend to produce a cooler Antarctic Ocean than currently exists. To appreciate why this is so, note that in summer, when ice is too extensive, the ocean gains less energy than in reality, due to the higher surface albedo. On the other hand, since there is too little ice in winter, the ocean is then exposed to unrealistically severe atmospheric cooling. (Although this also warms the atmosphere, the overall effect is a cooling of the system because of increased outgoing longwave radiation.) The implication of this is that the introduction of ice dynamics to achieve an improved seasonal cycle may accelerate the gradual erosion of the ice cover.

In the light of the discussion thus far, it is perhaps surprising that a fairly realistic seasonal cycle in ice extent should be obtained in the Northern Hemisphere (albeit with a sizable bias towards excessive ice). However the reasons for this are straightforward enough. First, the more enclosed nature of the Arctic reduces the importance of ice advection, so that the omission of ice dynamics is not quite as serious as it is for the Antarctic. Secondly, the thermal influence of the northern continents helps to maintain the contrast between summer and winter ice distributions in the model, just as it does in reality.

Another difference between the simulation of ice in the Arctic and the Antarctic is that the Arctic ice shows no sign of slowly decaying - indeed it is consistently too extensive. There appears to be more than one reason for this problem. The most obvious cause is again an indirect result of the insufficiently fine resolution of the ocean model grid. In order to avoid numerical problems, it is necessary to choose the horizontal momentum diffusion coefficient to be larger than in the real ocean, and this leads to western boundary currents that are too broad

and sluggish. For instance, present estimates of the peak mass transport of the Gulf Stream are of the order of 100 Sv, according to Niiler (1987), whereas the transport in the model fluctuates between about 20 Sv and 35 Sv. The meridional advection of heat by the boundary currents is similarly inadequate, and this must encourage the ice to spread further south than in the real world. As an example of the impact of this, consider the Barents Sea. Although perpetually icy in the model, it is largely ice-free in reality, mainly due to warm water advection by the ultimate extension of the North Atlantic Drift.

However, it appears that systematic errors in the atmospheric model also contribute to the over-abundance of Arctic ice. This emerges from consideration of the results of the 4AC experiment, in which the atmospheric model was run alone, with SSTs and ice extents prescribed from climatology. A particular problem with the 4AC integration was a very high incidence of low-level cloud over Canada and the USSR, especially during winter and spring. This blanket of cloud kept surface temperatures too high during winter, but delayed the onset of snowmelt by around a month in spring. Not only was summer thereby shortened, but surface temperatures were generally reduced, because the soil stayed damp and cloud cover persisted. In fact, summer surface temperatures were typically about 4°-5°C cooler than climatology in these areas. In the coupled model, this process forms part of a feedback loop, in which excessive sea-ice and low SSTs exacerbate the problems over land, and insufficient warm air advection from the continents in summer leads in turn to excessive ice. (One result of this is that surface temperatures over Canada and the USSR are cooler than climatology by about 10°-15°C in summer.)

This feedback spiral is given a further twist by a shortfall in the solar radiation absorbed at the surface of the Arctic Ocean in summer (ameliorated to some extent by the programming error mentioned above). This is due partly to the direct albedo feedback, but also to insufficient incident flux, values of which are typically less than half of the climatological estimates. This is again associated with low-level cloud, but the detailed reason for the error is different. In this case the model is quite correct to generate high amounts of cloud, because climatologies such as Volume 3 of the Soviet Navy Atlas (Gorshkov, 1983) and Volume 14 of the World Survey of Climatology (Orvig, 1970) agree that the Arctic Ocean certainly is very cloudy in summer. The trouble is that the model clouds have radiative properties more appropriate to clouds in lower latitudes, and so transmit too little solar radiation.

There is an interesting corollary to the foregoing analysis. Although systematic errors in both ocean and atmosphere models share the responsibility for the anomalously high amounts of Arctic ice, it seems that the atmospheric errors take effect mainly in summer, while the oceanic errors are if anything more important in winter, when ice is not confined to the Arctic proper, but spreads south into the North Pacific and North Atlantic. It would therefore be wrong to assume that, if improvements were made to one model but not the other, the resulting reduction in ice coverage would be uniform throughout the year.

Considering that the ice limits in the Northern Hemisphere are too extensive, it is surprising that the seasonal variation of ice concentration in the central Arctic is so realistic. One might well suspect that the programming error that enhances the solar radiation absorbed by leads actually improves the Arctic simulation in this respect. To test this idea, a short experiment was carried out in which the last July and August of the main integration were repeated with the error corrected. It was found that the lead area at the end of August was indeed significantly less than in the main integration.

Much of this discussion has focused on features of the ocean and atmosphere models rather than on the ice model itself. This is inevitable, given the highly interactive nature of the coupled system. However, it is strange that no clear evidence of deficiencies in the thermodynamics of the ice model has so far emerged. One can only conclude that errors in the forcing fluxes provided to the ice model are at present more significant than the weaknesses in the model itself, and may even be concealing them. The important exception to this statement is that the absence of ice dynamics does have a significant impact on the simulation, especially in the Antarctic.

5. FUTURE DEVELOPMENTS

As a result of the integration described in sections 2 and 3, we have been able to identify a number of areas of the coupled model where further work could produce useful improvements in the simulation of sea-ice, bringing closer the goal of using the coupled model for realistic experiments on climatic change. It is now appropriate to consider the prospects for such work.

There are good reasons for adding a representation of ice dynamics to the model. First, some of the problems with the simulation seem to result from its omission, as discussed in section 4. Secondly, there are also potential problems, which have not been observed in the current experiment but might become important in a longer one. These include the lack of polewards salt transport due to ice advection, which may well be important in the formation of the deep water of the world's oceans. (Washington & Parkinson, 1986.) Thirdly, Hibler (1984) concluded that the presence or absence of ice dynamics affected the sensitivity of his Antarctic model to an atmospheric warming, such as might be produced by climatic change. This suggests that experiments designed to understand climatic change need to include ice dynamics.

However, satisfactory models of ice dynamics, such as Hibler's (1979) model, seem to be quite expensive to run, mainly because of constraints imposed by the ice rheology. There is thus a need to strike a careful balance between achieving a realistic simulation on the one hand, and consuming a sensible proportion of the available computing resources on the other. Clearly there would be little to be gained from including ice dynamics at the price of having to degrade the representation of atmospheric radiation, for example.

Meanwhile, the possibilities of improving other aspects of the ice model should not be neglected. For instance, the next experiment with the coupled model will include a simple parametrization of the formation of so-called 'white' ice. This is ice that forms at the snow/ice interface when snow accumulates to such an extent that it depresses the ice below the waterline. Water then infiltrates the snow and tends to freeze there, forming white ice. Modelling studies by Ledley (1985) suggest that the inclusion of this process would be worthwhile. Refinement of the parametrization of surface albedo is also desirable, as noted in section 2. A useful step in this direction would be making the albedo depend on the existence of snow cover, and possibly on the depth of snow.

As far as the atmospheric model is concerned, a new benchmark version has been under development for some time, and a control integration of this has been carried out. The main new feature of this model that is relevant here is the use of cloud water content as an explicit prognostic variable. This enables the amount of cloud to be generated in a more physically-based manner than in the 4AC model, which computed cloudiness from an empirical formula involving the relative humidity in a grid box. The scheme has shown an encouraging improvement in the amounts of cloud over the northern continents in winter and spring, with associated improvements in the surface radiative fluxes (Smith, 1990). In the longer term, work on the use of interactive cloud optical properties may lead to further improvements in the surface radiative fluxes.

With regard to the ocean model, it appears from section 4 that major benefits could result from an increase in horizontal resolution. This would be unlikely to cure completely the problems noted there, but should considerably reduce their magnitude.

One process not represented in the current coupled model is the flow of fresh water from rivers into the oceans. It is possible that fresh water runoff from Eurasia and North America plays a significant role in the Arctic Ocean, by helping to maintain the cold halocline there. The halocline acts like a lid, preventing warmer but saltier Atlantic water below it from mixing with the surface water. There has been speculation that the large-scale diversion of Siberian rivers once contemplated in the USSR might result in the halocline becoming eroded, perhaps leading to the release at the surface of the heat stored in the Atlantic water, and thus to the destruction of the Arctic ice cover. (These issues have been reviewed by Cattle (1985).) Although the credibility of this scenario is an open question, it is interesting that the present model in effect performs the ultimate river diversion experiment, by eliminating the input from all the Arctic rivers. It would clearly be preferable to include river runoff as part of the standard model, with the option of diverting some or all of the Arctic rivers if one wanted to examine the consequences of doing so. (There are also other reasons for wishing to include river runoff, such as reducing the drift in the salinity field during long integrations of the ocean model.) Work on the incorporation of runoff into the coupled model is underway.

6. CONCLUSION

Sea-ice plays a significant role in the global climate system. The modelling of sea-ice, and its interactions with the atmosphere and the ocean, is a necessary ingredient of any serious attempt to simulate global climate. While the simulation of sea-ice in the coupled model is reasonable in some respects, several weaknesses are readily discernable. The causes of these are largely understood, and the prospects for improvement are good, although considerable effort will be necessary.

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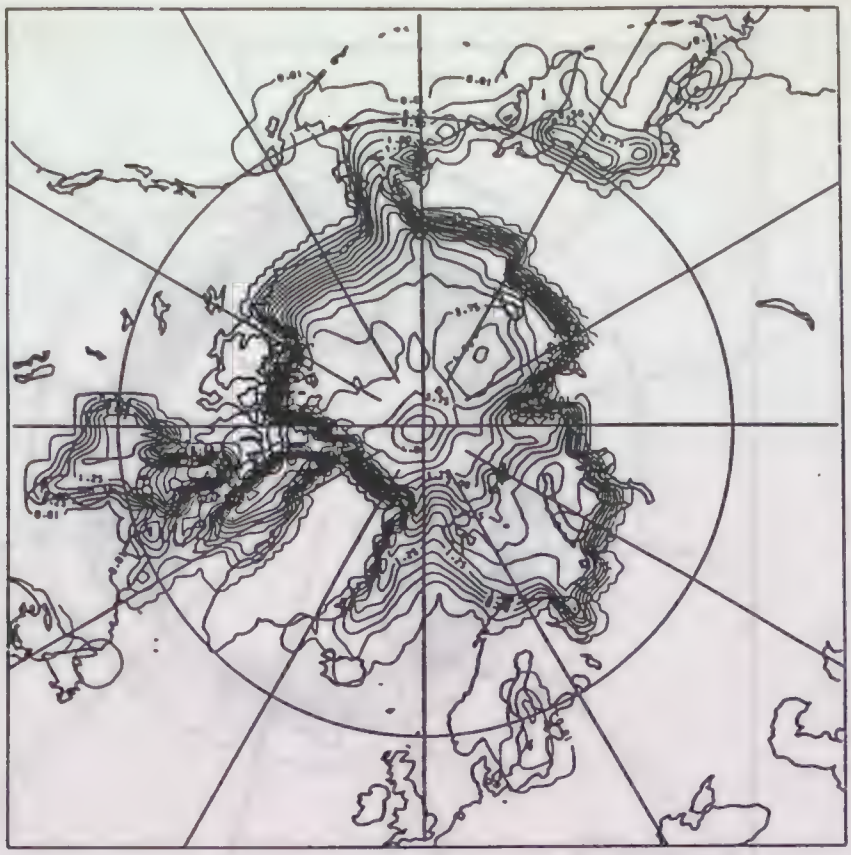

Figure 1. Simulated northern hemisphere sea ice thicknesses for March. The contour interval is 0.25m

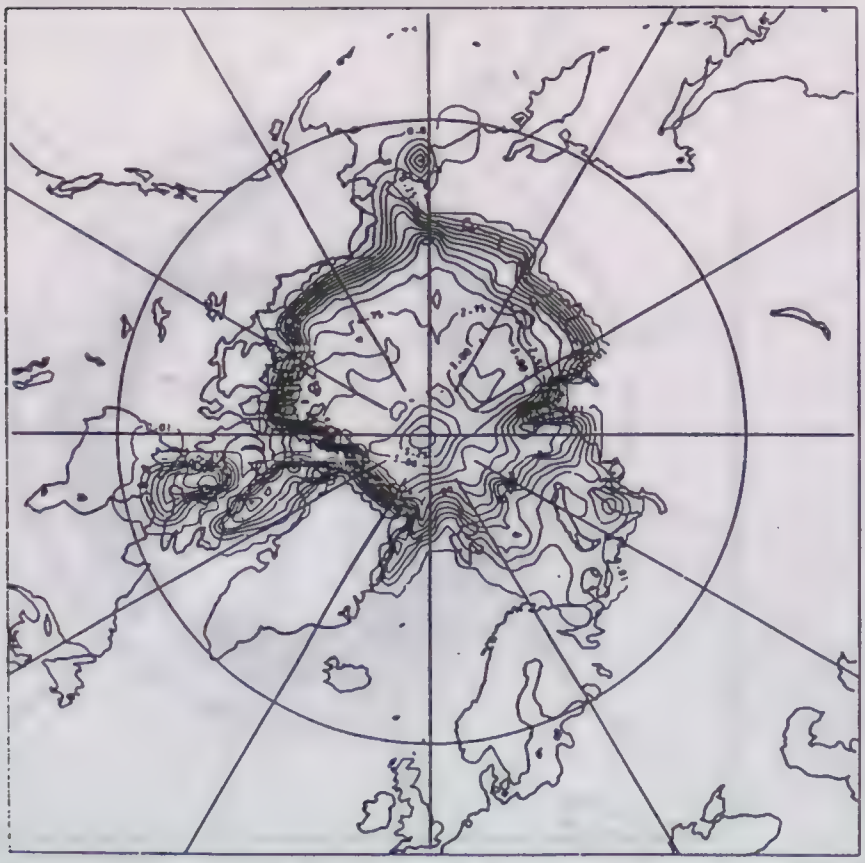

Figure 2 Simulated northern hemisphere sea ice thicknesses for September. The contour interval is 0.25m

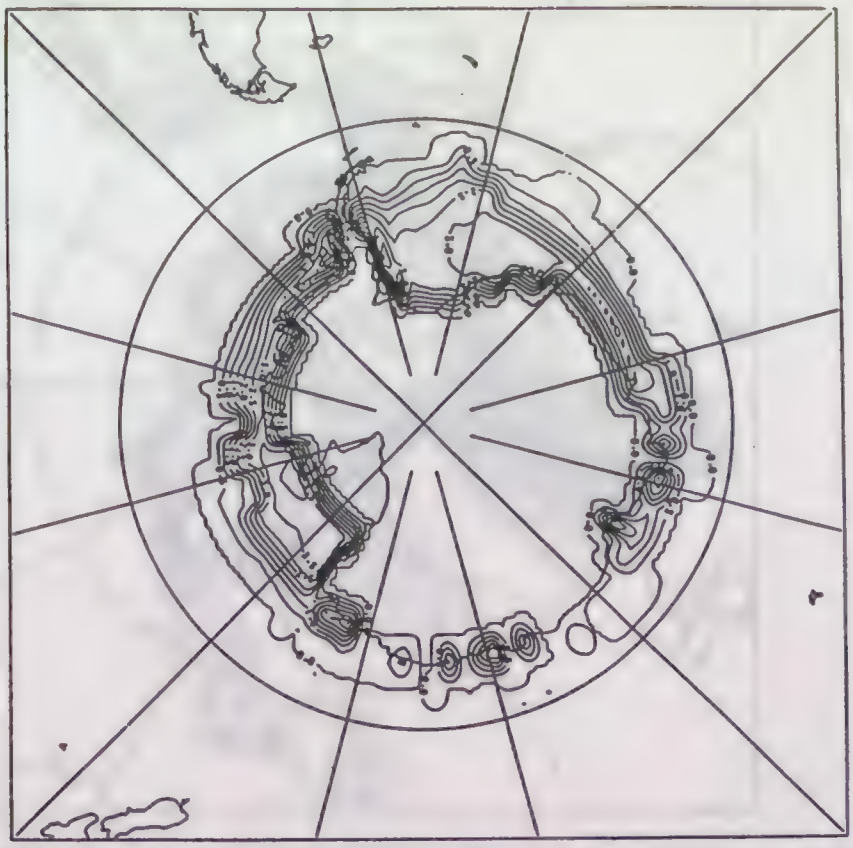

Figure 3 Simulated southern hemisphere sea ice thicknesses for February. The contour interval is 0.5m

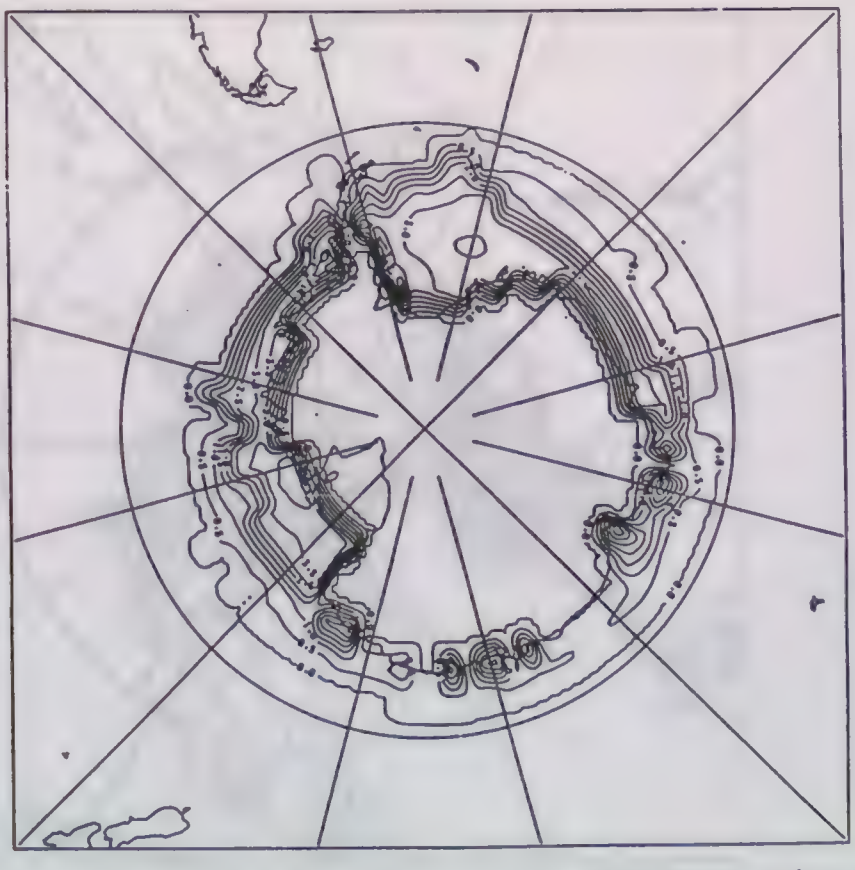

Figure 4 Simulated southern hemisphere sea ice thicknesses for October. The contour interval is 0.5m

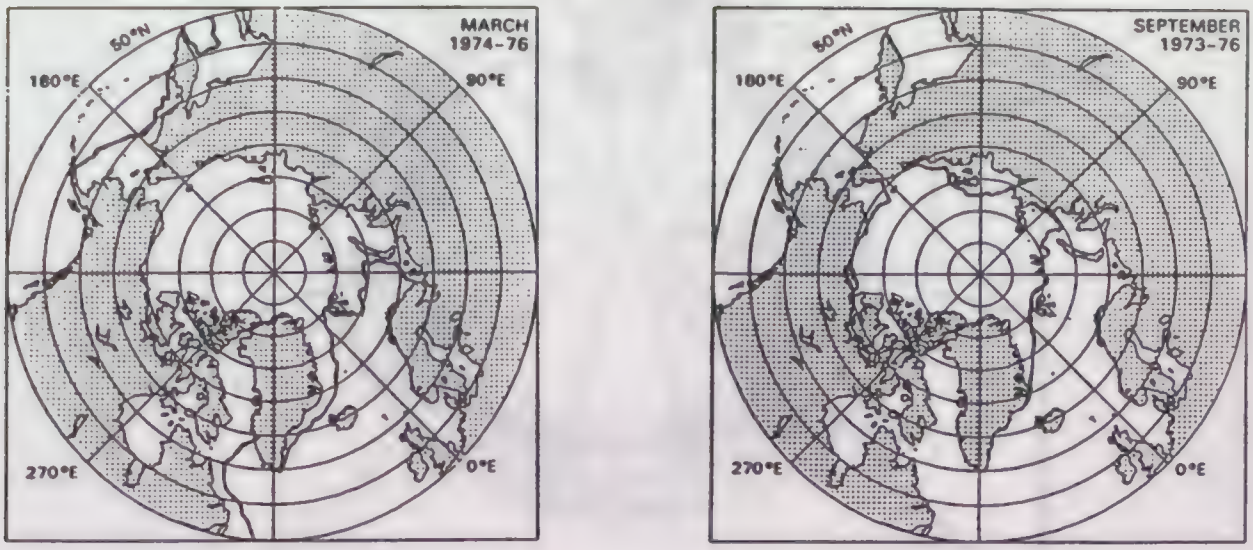

Figure 5 Northern hemisphere sea ice extents for March and September, derived from satellite data (see text).

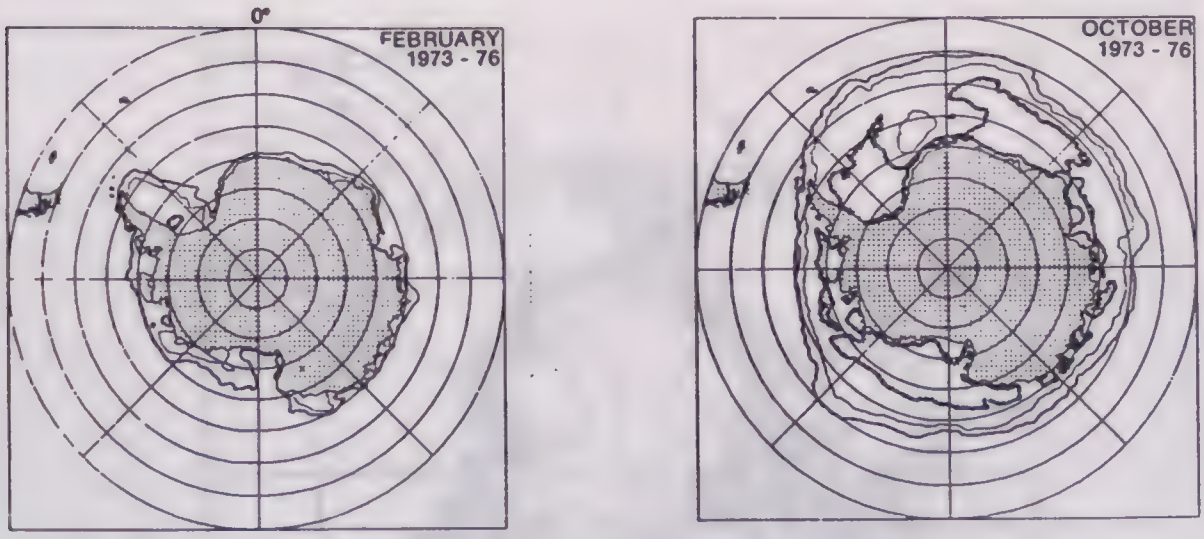

Figure 6 Sea ice extents for February and October, from satellite data. Here the solid, dotted and spiked lines indicate ice concentrations of 15%, 50% and 85% respectively.

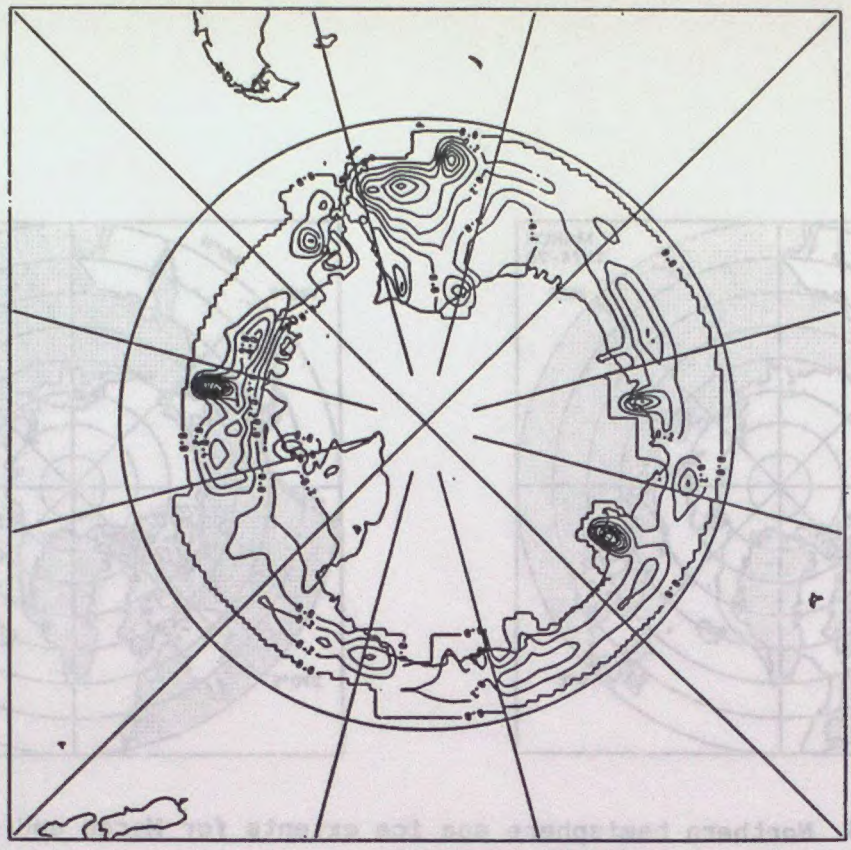

Figure 7 Mean simulated southern hemisphere ice thickness for February, year 3, minus the mean ice thickness for February, year 4. The contour interval is 0.2m.

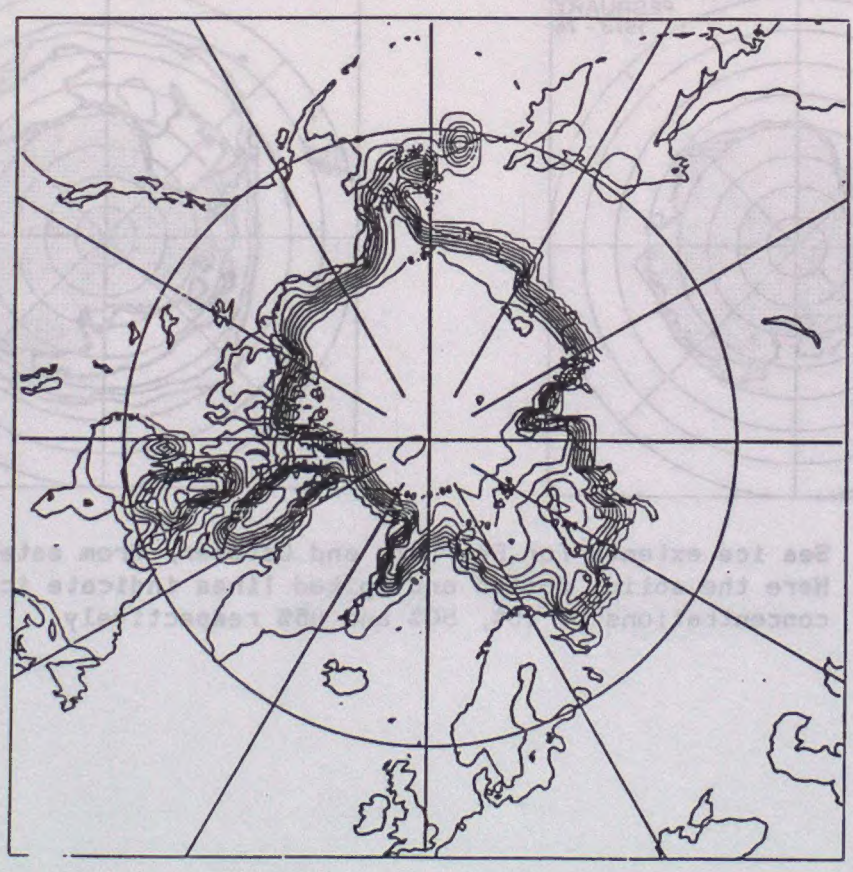

Figure 8 Simulated ice concentrations for August. Contours are drawn at 0.01, 0.1, 0.2, ..., 0.9.

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